

Paleoclimate Data Constraints on Climate Sensitivity: The Paleocalibration Method

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Abstract

The relationship between paleoclimates and the future climate, while not as simple as implied in the “paleoanalog” studies of Budyko and others, nevertheless provides sufficient constraints to broadly confirm the climate sensitivity range of theoretical models and perhaps eventually to narrow the model-derived uncertainties. We use a new technique called “paleocalibration” to calculate the ratio of temperature response to forcing on a global mean scale for three key intervals of Earth history. By examining surface conditions reconstructed from geologic data for the Last Glacial Maximum, the middle Cretaceous and the early Eocene, we can estimate the equilibrium climate sensitivity to radiative forcing changes for different extreme climates. We find that the ratios for these three periods, within error bounds, all lie in the range obtained from general circulation models, 2-5 K global warming for doubled atmospheric carbon dioxide. Paleocalibration thus provides a data-based confirmation of theoretically calculated climate sensitivity. However, when compared with paleodata on regional scales, the models show less agreement with data. For example, our GCM simulation of the early Eocene fails to obtain the temperature contrasts between the Equator and the Poles (and between land and ocean areas) indicated by the data, even though it agrees with the temperature data in the global average. Similar results have been reported by others for the Cretaceous and for the Last Glacial Maximum.

I. Introduction

Climate sensitivity can be defined as the eventual (or equilibrium) change in global mean surface temperature in response to a prescribed change in global mean radiative forcing. A conventional measure of climate sensitivity is the global warming ΔT_{2x} expected from doubling atmospheric CO_2 . Although this definition excludes time-dependent effects and regional details, it serves as a first approximation for approaching the issue of future global change. General circulation models of climate obtain ΔT_{2x} in the approximate range 2-5 K. Over the last two decades, no GCM with reasonable input assumptions has obtained a sensitivity much outside the 2-5 K range, but at the same time it has proved all too easy, by varying a model's assumptions within the bounds of plausibility, to move its sensitivity from one extreme end of the range to the other (e.g., Mitchell et al., 1989).

As an alternative to model-based predictions, Kellogg (1977) and Budyko and Izrael (1987) offered a "paleoanalog" approach based on direct analogy with past warm periods. The problem with this method, however, is that Twenty-First Century global warming would probably involve unprecedented rates of climatic change for which there are no satisfactory geologic analogs (Crowley, 1990). A less ambitious but more justifiable approach—"paleocalibration"—originated with Lorius et al. (1990). These authors used geologic data from the Last Glacial Maximum (LGM, 20 thousand years ago) to infer ΔT_{2x} without attempting to forecast the time-evolving behavior or regional details of the future climate. We extended the paleocalibration approach to the warm mid-Cretaceous era of 100 million years ago (Hoffert and Covey, 1992). In this paper we compare our paleocalibration results with more recent results of others, we discuss a fundamental criticism of the technique (Lindzen, 1993), and we

introduce a new paleocalibration data point, representing the early Eocene (55 million years ago).

Examination of paleoclimates with GCMs, the same models that predict future global warming, has been pursued for well over a decade (e.g., Hecht, 1985; Crowley and North, 1991). Although the paleocalibration technique is independent of GCMs, it leads naturally to questions of model fidelity in simulating paleoclimates. We conclude this paper with the results of new GCM simulations of the Eocene that address this point.

II. Method

In principle the paleocalibration technique is straightforward. For a given time interval, one obtains both the difference from present-day globally averaged surface temperature (ΔT) and the difference from the present-day globally averaged radiative forcing (ΔQ). ΔT is obtained by whatever geologic proxies are available (we will investigate different methods to estimate this value relative to what was used by Hoffert and Covey). ΔQ is derived by calculating or estimating the total effect of the heat trapped by greenhouse gases and the changes in absorption of solar radiation due to changes in solar luminosity, surface albedo and atmospheric aerosol content. The next and final step is simply to define the ratio $\Delta T / \Delta Q$ as the climate sensitivity, which is the global temperature response to the radiative forcing.

As an example, Hansen et al. (1993) estimated that for the Last Glacial Maximum, ΔT was -5 K and ΔQ was -7 W m^{-2} . Most of ΔQ arises from continental ice sheets and atmospheric aerosols reflecting more solar energy back to space (ice cores samples from the LGM clearly show that the atmosphere then contained much more dust than at present, though the exact amount it contained is controversial). A secondary term is the decreased trapping of infrared

radiation due to smaller atmospheric amounts of CO₂ and CH₄. The $\Delta T / \Delta Q$ ratio is 0.7 K (W m⁻²)⁻¹. This quantity can be converted to a value for comparison with global warming estimates by noting that a doubling of atmospheric CO₂ traps about 4 W m⁻² of infrared radiation. Then, using the values from Hansen et al., the expected global warming due to doubled CO₂ would be $(0.7) \times 4 = 3$ K, in the middle of the range of GCM estimates.

There are important limitations to the paleocalibration technique. First, the climate sensitivity as defined above says nothing about how long the system would take to respond to a given forcing. In the case of future global warming, the heat capacity of the oceans would introduce a lag time that predictions of future climate would need to account for by means other than paleocalibration. Note, however, that the primary factor determining the lag time is the equilibrium sensitivity value itself (Hansen et al., 1985; Wigley and Schlesinger, 1985). A second limitation of paleocalibration is that it determines only the globally averaged temperature response, not the pattern of regional response (nor the responses of other climatically important quantities like precipitation). As we discuss below, GCMs generally fail to simulate the observed difference in temperature response between Equator and Poles or between land and sea. Thus reliable forecasts of future regional climates are not yet obtainable from either paleocalibration or theoretical climate models.

A further subtlety comes in the definition of ΔQ . 4 W m⁻², for example, is the infrared trapping caused by doubled CO₂ *in the absence of other effects or feedbacks* such as changes in temperature, cloudiness or atmospheric water vapor content. ΔQ is defined as the total change in radiative flux at the top of the troposphere due only to changes in greenhouse gases, surface albedo, atmospheric aerosol content and solar luminosity. It may be thought of as the result of hypothetical, instantaneous changes in the above-mentioned factors,

before temperature, clouds or water vapor have a chance to respond. Such responses would of course occur simultaneously with changing radiative forcing in the real world, so ΔQ cannot be measured directly. For example, one cannot expect satellite observations to record a decrease of several W m^{-2} in the infrared flux of Earth to space as greenhouse gases increase. Instead one would expect atmospheric temperature to increase to restore an approximate balance of absorbed solar energy and emitted infrared (i.e., global warming due to an enhanced greenhouse effect). Despite its hypothetical nature, ΔQ is well defined; given a specified set of changes in greenhouse gases, aerosols and surface albedo, ΔQ can be found as a straightforward exercise in radiative transfer.

In short, paleocalibration defines the climatic feedbacks involved in cloud and water vapor changes as part of the response ΔT rather than part of the forcing ΔQ . The technique in effect measures the sum of cloud and water vapor feedbacks by observing ΔT . On the other hand, much slower processes like changes in atmospheric CO_2 and the growth and decay of continental ice sheets are included in ΔQ , the forcing. Distinguishing the fast feedbacks contained in ΔT from the forcing factors in ΔQ is thus a matter of scale separation. Paleocalibration does not aim to identify the root causes of past climatic changes, such as the causes of ice sheet growth and decay or of glacial-to-interglacial greenhouse gas variations. Instead the technique aims to measure the feedbacks that translate such root causes into temperature change. Feedbacks due to clouds and water vapor account for most of the uncertainty in the model estimates of future global warming.

III. Review of Results and a Fundamental Criticism

Table 1 compares our previous results (Hoffert and Covey, 1992) with subsequent estimates of Barron (1993) and Hansen et al. (1993). Note that we

estimated $\Delta T / \Delta Q$ for both the Cretaceous and the LGM, whereas Barron dealt with the Cretaceous only, and Hansen et al. dealt with the LGM only. Both our Cretaceous and LGM estimates gave $\Delta T_{2x} \sim 2$ K, at the low end of the GCM prediction range. On the other hand, the LGM estimate of Hansen et al. gave $\Delta T_{2x} \sim 3$ K, near the center of the GCM range, and the Cretaceous estimate of Barron gave $\Delta T_{2x} \sim 5$ K, at the upper end of the range. The difference between our estimate and that of Hansen et al. arises from differing values of ΔT we used for the LGM. We used $\Delta T \approx -3$ K, obtained by taking a global average of the sea surface temperatures compiled by the CLIMAP analysis of LGM data. For land areas we simply assumed that the LGM cooling was identical to that of ocean areas in the same latitude zones. This assumption was based on the general principle that the atmosphere efficiently smooths out temperature contrasts between land and sea within each latitude zone. However, terrestrial geologic data suggest that the Ice Age cooling over land was significantly larger than that over ocean, and it has even been suggested that the ocean ΔT values obtained by CLIMAP are too small in magnitude (Rind and Peteet, 1985; Guilderson et al., 1994). Based on such considerations, Hansen et al. chose $\Delta T \approx -5$ K for the LGM and obtained a correspondingly higher estimate of ΔT_{2x} than ours.

Table 1: Intercomparison of Paleocalibration Estimates

	ΔT [K]	ΔQ [W m ⁻²]	$\Delta T_{2 \times CO_2}$ [K]
Hoffert and Covey (1992) LGM	-3 ± 0.6	-6.7 ± 0.9	2 ± 0.5
Hansen et al. (1993) LGM	-5 ± 1	-7.1 ± 1.5	3 ± 1
Hoffert and Covey (1992) Cretaceous	9 ± 2	15.7 ± 6.8	2.5 ± 1.2
Barron (1993) Cretaceous	9 ± 3	8 ± 3.5	~ 5

Barron's discussion of ΔT_{2x} provides a further gauge of uncertainty in paleocalibration, although it was not presented as such. Barron's presentation was in terms of (a) the amount of Cretaceous atmospheric CO_2 required to match different possible values of ΔT_{2x} and (b) likely values for Cretaceous CO_2 . Barron used essentially the same estimates of Cretaceous ΔT as we did (in fact we used his previously published estimates) and arrived at 9 K global mean warming. Assuming logarithmic scaling of CO_2 radiative forcing and a 2-5 K range of GCM sensitivity to CO_2 doubling, the implied Cretaceous CO_2 amounts range from 3 to 23 times present. Considering 2-6 times present atmospheric CO_2 as a range of likely Cretaceous CO_2 amounts, Barron inferred that the climate's true sensitivity must lie in the upper range of model results in order to bring the implied Cretaceous CO_2 amounts within reasonable bounds. In effect this exercise is paleocalibration of climate sensitivity using $\Delta T \approx 9$ K and calculating ΔQ only from a 2-6 fold increase in CO_2 . We used 2-11 times present CO_2 (perhaps an overestimate of the possible range) and we also included surface albedo and solar luminosity changes not considered in Barron's calculation. As a result our ΔQ was twice as large as his, and thus our inferred ΔT_{2x} was only half as large.

It should be noted that a recent revision of Cretaceous temperatures by Sellwood et al. (1994) obtained "minimum estimates" somewhat cooler than the Barron's lower limits. Sellwood et al., however, failed to consider the substantial Equator-to-Pole gradient of oxygen isotope ratio (their proxy for temperature). Properly taking this gradient into account substantially increases the tropical temperatures inferred from oxygen isotopes (Zachos et al., 1994; Hoffert et al., submitted).

The most important conclusion from Table 1 is that paleocalibration gives roughly the same range of possible values for ΔT_{2x} as GCMs do. Although it does not change the conventional wisdom about the magnitude of potential human-induced climatic changes, paleocalibration strengthens the GCM-based theory by providing independent confirmation. Of course it would be useful if paleodata could be used to narrow the range of uncertainty in ΔT_{2x} . Our own previous results (Hoffert and Covey, 1992) and the preliminary Eocene analysis given in the following section imply that ΔT_{2x} lies at the low end of GCM predictions. As discussed above, however, alternate interpretations of the paleodata can push ΔT_{2x} upward. We must admit that our estimate of 3 K for Ice Age cooling is smaller than the consensus value among paleoclimatologists, and that our ± 0.6 K LGM error limits accounted only for the scatter of CLIMAP longitude-averaged sea surface temperatures relative to a smooth curve, not the range of different interpretations of the paleodata.

In addition to controversy over the most appropriate input values for the paleocalibration technique, there is a fundamental objection to the technique itself. Paleocalibration makes the basic assumption that globally averaged temperature response depends on the globally averaged forcing, i.e., that ΔT is a unique function of ΔQ . Lindzen (1993) has asserted to the contrary that under “an altered *distribution* of heating . . . major changes in global climate may occur, even if the sensitivity to changing CO_2 is extremely small [emphasis added].” The coming and going of Ice Ages, for example, are clearly associated with small changes in the distribution—but not the global mean—of insolation, caused in turn by small variations in Earth’s orbit about the Sun (Milankovitch forcing; see Imbrie and Imbrie, 1979). Also, the glacial-interglacial CO_2 variations which comprise the main part of Ice Age ΔQ may themselves be caused by glacial-interglacial climate changes. So how can one infer climate sensitivity from them?

The answer is that paleocalibration does not attempt to identify root causes of climatic change, only to measure the feedbacks that determine the level of response to those root causes and result in the observed paleotemperatures. Thus, as noted in our brief exchange with Lindzen (Hoffert and Covey, 1993), Milankovitch forcing may well modulate Ice Ages, but it is difficult to explain 3-5 K global cooling without invoking positive feedbacks that amplify the climate's response to small changes in the distribution of insolation, and that also would produce significant global warming from doubled CO₂. An attempt by Kirk-Davidoff and Lindzen (1993) to do so illustrates the difficulty. They presented a simple climate model in which significant global mean temperature changes resulted from merely changing the transport of heat from the Equator toward the Poles. In principle this result is not surprising, because nonlinear feedbacks can create a situation in which moving heat from location to another will change the global mean temperature (Robock, 1978). To obtain significant ΔT , however, Kirk-Davidoff and Lindzen assumed negative water vapor feedback in the tropics some 30 times greater in magnitude, and of opposite sign, to the feedback conventionally derived from satellite observations (Warren and Schneider, 1979; Rind et al., 1991).

We find it difficult to imagine how the negative water vapor feedback assumed by Kirk-Davidoff and Lindzen could be reconciled with the satellite data. The point made by Sun and Lindzen (1993), that "the water vapor content of the air above the trade inversion over the subtropics is not directly related to the sea surface temperature immediately below," is certainly reasonable. Nevertheless the satellite data appear to argue strongly for an indirect effect amounting to positive water vapor greenhouse feedback. A $2 \text{ W m}^{-2} \text{ K}^{-1}$ slope of outgoing longwave radiation with temperature appears not only in spatial correlations but also in examination of different times at the same location (e.g.,

Figure 2 in Raval et al., 1994). This value of the slope corresponds to $\Delta T_{2x} \sim 2$ K (without taking albedo feedback into account), which could not be attained without substantial positive water vapor feedback.

IV. Paleocalibration for the Eocene Earth

Our claims would be put on a firmer foundation if they could be confirmed with more data points representing additional time intervals of Earth history. Here we add a third data point, representing an interval of Earth history that is located temporally between the two established points of the LGM and Cretaceous.

The early Eocene climate was probably the warmest since the Cretaceous (e.g., Shackleton and Boersma, 1981; Wolfe, 1985; Barron, 1987; Crowley and North, 1991; Cerling, 1992; Sloan and Barron, 1992). Surface geography for the early Eocene was similar to today's with the following differences: Australia was in a more southerly position than today, located adjacent to Antarctica; India was located in the tropics and had not yet collided with Asia; the Mediterranean was larger than it currently is; the Himalayas were not as great in elevation; also the Rockies, Andes, and Transantarctic Mountains may have been somewhat reduced in elevation. The polar regions had little or no ice and subtropical plants existed within the Arctic Circle. On the other hand tropical temperatures were similar to today's values (Zachos et al., 1994). As shown below (Figure 1), the Eocene appears to be an extreme case in which tropical temperatures were no warmer, and perhaps even colder, than at present despite significant global mean warming. If any era exemplifies Lindzen's theory of global change through changes in Equator-to-Pole heat distribution, the Eocene should.

Solar output 55 million years ago was similar to that of the present day within uncertainty bounds. Surface albedo would have been less than present

during the Eocene due to (1) lack of most, if not all, perennial continental and sea ice, (2) higher sea levels, leading to greater areal coverage of oceans, (3) presence of deciduous forests at high latitudes of both hemispheres and (4) lack of extensive deserts. Estimates of atmospheric CO₂ during the early Eocene come from theoretical models of Berner (1991; 2 x preindustrial) and from geochemical interpretations of organic carbon (Freeman and Hayes, 1992: 2x; Arthur et al., 1992: 3-6x) and soil nodules (Cerling, 1992: 2x). These estimates agree that CO₂ was higher than present, but they vary over a broad range of 2 to 6 times preindustrial values. There is some thought that methane concentrations may have been higher than present during the Eocene due to extensive areas of swamps and wetlands (Sloan et al., 1992), but there is no direct evidence for this. There is no evidence whatsoever regarding atmospheric aerosols during this time.

Increased atmospheric carbon dioxide is probably the dominant term in Eocene radiative forcing compared with the present. For the contribution to ΔQ from CO₂ we take the full range of estimates discussed above, 2-6 times the preindustrial value. Using logarithmic scaling from 4.4 W m⁻² for CO₂ doubling, this gives a contribution of 7.9 ± 3.5 W m⁻². We neglect possible contributions from methane. Of the four above-mentioned factors that contribute to a change in surface albedo, lack of ice probably dominates forest growth and lack of deserts (Bonan et al., 1992). Covey et al. (1991) estimated that 2 to 3 W m⁻² radiative forcing would result from the complete disappearance of sea ice from the present-day Earth, but we have excluded sea ice changes from our definition of ΔQ because we want to include such changes in the feedback processes measured by the paleocalibration technique. Accordingly we consider only the remaining contributor to changes in surface albedo, namely higher sea levels and the resulting decrease in the fraction of relatively high-albedo land areas. As a

crude first estimate we set ΔQ due to this effect at two-thirds the Hoffert and Covey (1992) estimate of Cretaceous ΔQ due to surface albedo effects. The result is $3.9 \pm 0.6 \text{ W m}^{-2}$. Combining this with the contribution from increased CO_2 (and taking the square root of the summed squared error estimates, as is appropriate for independent sources of error) gives a total ΔQ of $11.8 \pm 3.6 \text{ W m}^{-2}$.

Early Eocene paleoclimate data indicate a world that was warmer than present, with greatest warming at high latitudes and little or no warming at low latitudes. Figure 1 shows the difference between mean annual Eocene and present surface temperatures as a function of latitude. We simply plotted local ΔT inferred from the paleodata at each available point on the globe together with the paleolatitude of each point (data taken from Sloan and Barron, 1992, and Zachos et al., 1994). We then fit a fourth-order polynomial in the sine of latitude to the data, weighting all points equally. The equal-weighting assumption is of course a crude approximation, but the data are so sparse that a more sophisticated treatment, such as interpolation in latitude-longitude space, seems unjustified to us. The globally averaged temperature change obtained from the integral of the fitted curve is $\Delta T = 4.3 \text{ K}$. (Note that both land and ocean points are used to obtain this value; if land points are excluded the same procedure gives $\Delta T = 3.3 \text{ K}$.) To obtain an error estimate for this figure we first note that the root-mean-square scatter of points about the fitted curve is 2.6 K . This should be divided by the square root of the number of data points (29) to obtain the contribution to uncertainty in the global average: $\pm 0.5 \text{ K}$. We must also include errors in translating proxy measurements to temperatures. Although these are more difficult to quantify, there appears to be consensus among those working with the ocean data that $\pm 2 \text{ K}$ is a reasonable estimate for the total error in this category. Errors in the land data may be higher, but we will use $\pm 2 \text{ K}$ because

most of the data in Figure 1 come from the ocean (we do not divide 2 K by $\sqrt{29}$ because the errors in proxy-to-temperature conversion may well be systematic). Combining ± 0.5 K and ± 2 K in root-sum-square fashion then gives a final estimate of $\Delta T = 4.3 \pm 2.1$ K.

In Figure 2 our Eocene estimates of ΔT and ΔQ are shown together with our earlier LGM and Cretaceous estimates (Hoffert and Covey, 1992), and the LGM estimate of Hansen et al. (1993). The figure illustrates application of the paleocalibration technique by representing the data in graphical terms, as a plot of ΔT versus ΔQ for several climate states in Earth history. By definition the present-day climate is a point at the origin. The three past climates, Eocene, Cretaceous and LGM, provide four more points (counting the independent LGM studies by us and by Hansen et al.). Conventional wisdom about Earth's climate sensitivity predicts that all points should lie in the range obtained from general circulation models, corresponding to 2-5 K warming for a doubling of atmospheric CO_2 . Within error limits, the points do indeed lie within the theoretically predicted range. The Eocene data, however, stand out as implying a significantly smaller climate sensitivity than the Cretaceous or LGM data. Using Equation (4) in Hoffert and Covey (1992), the Eocene ΔQ and ΔT values presented above imply $\Delta T_{2x} = 1.6 \pm 0.9$ K. This result must be viewed with caution in light of the preliminary nature of our Eocene numbers.

The most robust conclusion from all the paleocalibration results considered together is that the observed data lie approximately within the range of climate sensitivity predicted by theoretical models (with a bias toward the low end of the range). A radical challenge to the GCM-based conventional wisdom, such as a claim that models overestimate ΔT_{2x} by an order of magnitude or more, would need to explain why the paleodata points in Figure 2 lie close to the model-predicted range. Correlation of course does not imply causation, but we

would find it surprising if the actual causes of global mean temperature change involved the distribution rather than the global mean of radiative forcing, as suggested by Lindzen (1993). In that case the correlation of data points in Figure 2 would be due to global mean climate changes causing changes in global mean radiative forcing (just coincidentally with the $\Delta T / \Delta Q$ ratio predicted by climate models). But the correlation of the four data points we have compiled—those of Hoffert and Covey (1992), the Eocene point discussed above, and the present-day climate by definition at the origin—is clearly positive ($r = 0.975$; $P < 0.03$). A negative correlation would be expected if the figure were revealing ΔT as the primary cause of ΔQ through the long-term carbon cycle. In that case increased temperature would lead to increased weathering and hence enhanced removal of CO_2 from the atmosphere, leading to decreased ΔQ .

To reduce error limits and gain more confidence in the placement of data points in Figure 2, we need a more thorough examination of the data for all time intervals plotted. For the Eocene this is especially intriguing because of newly available Russian data for the Eurasian continent that imply a much larger value for ΔT than Western reconstructions indicate. While our compilation of mean annual temperature data produces an Eocene global warming of 4.3 K over present values, a Russian compilation indicates mean global Eocene warming of 9.7 K (Hoffert, 1993). The Russian data span the entire Eocene epoch while our compilation is restricted to the early Eocene (see Sloan and Barron, 1992). However, since the early Eocene is thought to have been the warmest interval, the Russian estimates should have a cold bias and not a warm bias in comparison to our data. The discrepancy between the Eocene temperature estimates is an issue that will have to be clarified in the future.

V. Regional Climate Sensitivity

The foregoing conclusion, that GCMs and paleodata are in rough agreement, generally applies only to the global average of temperature. Considering regional scales in addition to global-mean scales, we note that a GCM simulation of the early Eocene, while obtaining globally averaged ΔT consistent with the data, fails to obtain the sharp equator-to-pole surface ΔT gradient and the proper land-sea thermal contrast that the paleoclimate data suggest (Sloan et al., in press; Sloan and Rea, submitted). The Eocene GCM results for 1, 2, and 6 times present atmospheric CO_2 show global warming relative to the control simulation of 1.0, 3.1, and 6.3 K respectively, consistent with the paleodata's implication of about 4 K warming under 2-6 times preindustrial CO_2 . Figure 3 shows the annual mean, longitude-averaged surface temperature increase over the present day for the $2 \times \text{CO}_2$ and $6 \times \text{CO}_2$ Eocene simulations. Also shown are the data points (same points as in Figure 1). It is evident from the figure that although the model's change in temperature approximately agrees with the data in the global mean, the distribution of ΔT with latitude obtained by the model is far too uniform. Put another way, the model obtains nearly the same sharp Equator-to-Pole contrast in absolute T for the Eocene as for the present day, whereas the geologic data imply that this temperature contrast was greatly reduced. For example, in mid-continental winters the model obtains below-freezing temperatures while data such as alligator and crocodile fossils clearly shows that these areas did not undergo seasonal extremes during the Eocene (Sloan and Barron, 1992; Markwick, 1994; Sloan, in press). Similar problems were encountered in Cretaceous simulations by Barron and Washington (1984) and in LGM simulations by Manabe and Broccoli (1985). The tendency of current GCMs to predict relatively uniform

global warming, including substantial tropical warming, in the face of observations that indicate that tropical surface temperatures have changed little in the geologic past is a key criticism of the reliability of these models (e.g., Horrell, 1990; Crowley, 1991; Lindzen, 1993).

We are left with a great irony in our effort to understand climatic change. When compared with paleodata, general circulation models show fair performance in the global mean but poor performance at the next levels of approximation, i.e., Equator-Pole and land-sea temperature contrasts. In fact the GCMs agree with each other less and less as the spatial scale of comparison is decreased (Grotch and MacCracken, 1991). There are important exceptions to this discouraging trend (COHMAP, 1988), but in general the predictions of GCMs, which give the most detailed simulations of climate available, are not reliable unless they are averaged to a global mean. Of course it is the regional details that matter to humans and to natural ecosystems. We can only hope that more rigorous examination of the inner workings of GCMs (e.g., Gates, 1992) will improve this situation.

VI. The Future of Paleocalibration

A glance at Figure 2 shows that even though there is rough agreement between GCMs and paleodata, both the range of the model results and scatter and error bars in the data are large. Uncertainties in the data must be reduced if the data is to distinguish among differing model results, rather than simply confirm that the climate's sensitivity lies approximately within the range of estimates from different models. Reducing uncertainties in the data would also test the validity of the paleocalibration technique itself. If a version of Figure 2 with more careful placement of data points and smaller error bars shows the points lying securely on a line through the origin, then the case for interpolating

to find ΔT_{2x} would be compelling. If on the other hand the correlation in ΔT - ΔQ space were to disappear, the technique would clearly fail.

There are two ways to narrow uncertainty in the paleodata. First, we can try to refine estimates for the time periods we have already considered. This task would involve examining new data, such as the Russian estimates discussed above. It should also involve treating the data in a consistent way. The estimates shown in Figure 2 were arrived at by different means. For example, our error bounds for Cretaceous ΔQ included the full range of published estimates, but our error bounds for Ice Age ΔT accounted only for the scatter of points in the CLIMAP sea surface temperature data, not the widely held view that the CLIMAP data set systematically underestimates the globally averaged ΔT . One way to impose consistency and completeness in estimating ΔQ , incidentally, would be to use a GCM as a single-step calculator of radiative forcing in the absence of feedbacks rather than time-integrating predictor of climate (Covey et al., 1991).

A second way to reduce paleodata uncertainty is to use more data points. Considering additional time periods for paleocalibration is especially attractive to us because it provides a test of the main assumption underlying the technique itself, namely the conventional wisdom that global mean temperature is (to first approximation) a unique function of global mean forcing. For example, in the Pliocene era 3 million years ago, globally averaged surface temperature was perhaps 1-3 K warmer than present and, until recently, atmospheric CO_2 was assumed to be about twice present levels (Crowley, 1991; Webb et al., 1993). More recent data, however, suggests that CO_2 levels may have been only 30-50% above preindustrial (Raymo and Rau, submitted), corresponding to a ΔQ of 1.7-2.6 W m^{-2} . For $\Delta T = 2$ K the implied climate ΔT_{2x} is 3.4-5.3 K, assuming as a crude first approximation that CO_2 is the only factor in Pliocene ΔQ .

In conclusion, despite the limitations of paleocalibration in providing accurate and regionally relevant connections between past and future climates, we believe the technique is promising. One point is clear in any case. Paleoclimates give us the only real-world data that includes global changes of the magnitude predicted to occur as a result of human perturbation of the atmosphere during the next century. Except for waiting for such changes to occur, examination of paleoclimatic data is the only way to directly test the validity of the models that predict such changes.

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Figure Captions

Figure 1: Paleotemperature data vs. paleolatitude for the early Eocene . ΔT is the difference between Eocene and present-day surface temperature at each point; the dashed line is a least-squares fit to the data points. Circles show marine data (Zachos et al., 1994) triangles show continental data (Sloan and Barron, 1992), and the “x” denotes a single marine data point excluded from the analysis because it occurs in a modern-day upwelling zone whose anomalously cool temperature overestimates ΔT .

Figure 2: Globally averaged ΔT vs. globally averaged ΔQ observed for several different paleoclimates. Shown for comparison is a range of climate sensitivity values obtained by theoretical models, equivalent to 2-5 K for CO_2 doubling.

Figure 3: Eocene paleodata (as in Figure 1) compared with two GCM simulations of the Eocene which assumed 2 and 6 times present levels of atmospheric CO_2 .







